Sequence stratigraphy, correlations between Wopmay Orogen and Kilohigok Basin, and further investigations of the Bear Creek Group (Goulburn Supergroup), District of Mackenzie, N.W.T.

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Abstract

Results indicate that the Rifle, Beechey, Link, and basal Burnside Formations are correlative with the lower member of the Odjick Formation (Coronation Supergroup). The lower Burnside Formation is also correlative with the middle member of the Odjick Formation. Correlatives of the Hackett Formation and Kimerot Group are not present in the Coronation Supergroup. In the Tinney Hills area the Rifle Formation is divided into four sequences.

The marine to alluvial transition in the north Tinney Hills is characterized by three main associations of facies which represent storm-influenced marine shelf, lower delta slope, and upper delta slope. Additionally, studies of areally extensive conglomerate intervals indicate transport of gravel across the entire Slave craton, in excess of 200 km. This requires a fundamental change in the distribution of subsidence across the basin. Areally-extensive conglomerates indicate reduced subsidence rates in the proximal part of the basin. The transition from lower Burnside Formation deltaic and distal alluvial facies to gravelly proximal alluvial facies probably records a shift from subsidence-dominated foreland sedimentation to erosion- and uplift-dominated sediment redistribution.

Résumé

Les résultats indiquent que les formations de Rifle, de Beechey et de Link ainsi que la formation de base de Burnside sont en corrélation avec le membre inférieur de la formation d'Odlick (supergroupe de Coronation). La partie inférieure de la formation de Burnside est également en corrélation avec le membre central de la formation d'Odjick. Des indications de corrélation entre la formation de Hackett et le groupe de Kimerot ne sont pas présentes dans le supergroupe de Coronation.

La transition des dépôts marins aux dépôts alluviaux dans la partie septentrionale des collines Tinney est caractérisée par trois grandes associations de faciès représentant la plate-forme marine influencée par les tempêtes, le talus inférieur du delta et le talus supérieur du delta. De plus, des études d'intervalles étendus de conglomérat indiquent qu'il y a eu transport de gravier en travers de tout le craton des Esclaves, soit sur plus de 200 km. Cela exige une modification fondamentale de la répartition de la subsidence dans l'ensemble du bassin. Les conglomérats occupant une grande superficie indiquent des taux de subsidence réduits dans la partie proximale du bassin. La transition des faciès deltaïques et alluviaux distaux de la partie inférieure de la formation de Burnside aux faciès alluviaux proximaux graveleux marque probablement une évolution de la sédimentation d'avant-pays dominée par la subsidence en une redistribution des sédiments dominée par l'érosion et le soulèvement.

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INTRODUCTION

Several projects were finished during the field season of 1988. These are discussed on a topical basis and include: 1) correlation of stratigraphic units in the Bear Creek Group, as determined on both a formational and sequence stratigraphic basis, from Kilohigok Basin to the autochthon of Wopmay Orogen (JPG); 2) identification and correlation of sequence stratigraphic units in the Rifle Formation(JPG, RDA, PM); and 3) description and interpretation of the marine-alluvial facies transition in the Burnside Formation, and stratigraphic relations within its lower part (DSM).

STRATIGRAPHIC CORRELATIONS BETWEEN KILOHIGOK BASIN AND WOPMAY OROGEN

A complete transect of the Bear Creek Group has been mapped from the Bear Creek Hills to the northern Tinney Hills along the east side of Bathurst Inlet, and from the Western River to Contwoyto Lake. These two outcrop belts are separated by the Bathurst Fault which has approximately 135 km of sinistral slip (Fig. 1). In addition to the mapping, sections were measured along the two transects, as well as at Rockinghorse outlier and in the autochthon of Wopmay Orogen. Figure 2 shows the correlations (palinspastic) between the Bear Creek Group and the lower Odjick Formation in the autochthon of Wopmay Orogen at the north end of Takijuq Lake. The 15 sections that form the basis for this figure were measured on a bed-by-bed basis, and then condensed to generate summary sections that illustrate first-order stratigraphic features.

All stratigraphic units thicken toward the southeast (palinspastic) corner of the basin (Fig. 2). Section 15 is located at the north end of the Tinney Hills and it should be noted that equivalent sections at the southern Tinney Hills and Bear Creek Hills are approximately twice as thick, and therefore not plottable at the scale chosen for this paper. For the most part, formation boundaries are equivalent to sequence boundaries, and therefore closely approximate time slices (sensu Van Wagoner et al., 1987)

Across the Slave craton, all units thin over the Gordon Bay Arch, culminating in major downcutting of the Burnside Formation fluvial units, and thicken again into the Wolverine Canyon area (Fig. 2). This indicates that the Gordon Bay arch was a continually active feature throughout all of Bear Creek Group time. Continuing to the west, the lowermost units (e.g. Hackett Formation) thicken toward the Rockinghorse area, intermediate units (e.g. Beechey and Link formations) show no significant changes in thickness, and uppermost units (e.g. Burnside Formation) thin significantly and cut down into older units. This suggests that this area evolved from intial low subsidence, to no net subsidence, to a positive arch area later on. Toward the Takijuq area, older units (e.g. correlatives of Hackett Formation) were initially not deposited or have been removed by arching of this area during deposition of the equivalents of the lower Rifle Formation. Uppermost units (e.g. probable upper Rifle through Burnside equivalents) thicken markedly relative to the Rockinghorse area. This suggests that the Takijuq area initially was a positive arch area (Hackett and lowermost Rifle time) evolving to a rapidly subsiding area with time (upper Rifle through Burnside time).

Several statements can be made concerning the correlations between Wopmay Orogen and Kilohigok Basin. First, equivalents of the Hackett Formation are missing in the Takijuq Lake area (Fig. 2). Second, the base of the Coronation Supergroup in the Takijuq Lake area is almost coincident with the condensed interval of the lower Rifle Formation. This is probably the best time line between basins and allows a high degree of confidence to be assigned to these correlations. Third, the middle Odjick Formation is almost certainly correlative with the lower Burnside Formation.

Given the correlations presented in Figure 2 and the several ash beds which have been found at critical stratigraphic levels (Fig. 2), it should be possible to erect a calibrated chronostratigraphy between the lower Coronation Supergroup and Bear Creek Group. Inital results of U-Pb zircon dating on ash beds suggest that the age of the lower Hackett Formation (section 15; Fig. 2) may be 1.95-1.97 Ga (S.A. Bowring, pers. comm., 1988), and that the Link Formation (section 6; Fig. 2) is on the order of 1.93 Ga (Bowring and Grotzinger, in press). An ash bed collected at section 12 (Fig. 2) lies on the contact between the upper Rifle and Beechey formations, and although not yet dated, it should be on the order of 1.94-1.95 Ga. If correct, then this would potentially date the timing of onset of passive margin subsidence in Wopmay Orogen. Note that this boundary continues into Wopmay Orogen (section 1), where it most likely correlates with the top of a coarsening-

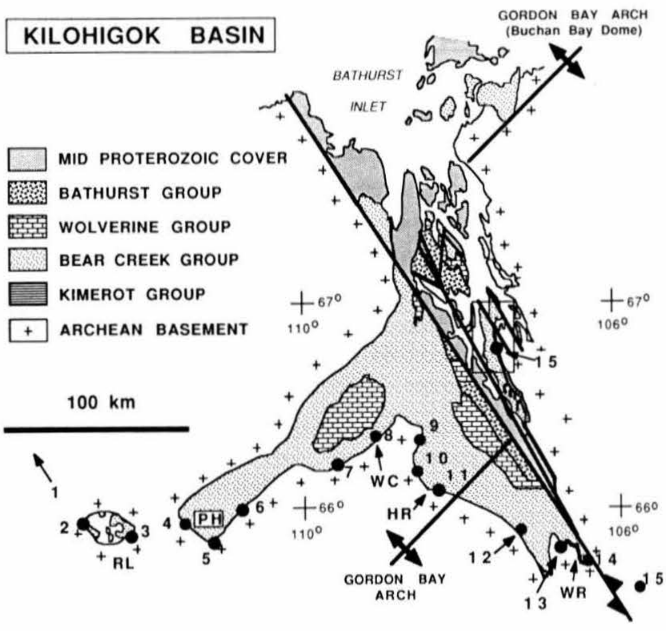

Figure 1. Location map of measured sections shown in Figure 2. Kilohigok Basin is shown without palinspastic restoration for movement along the Bathurst Fault. 15* marks the palinspastic location of section 15 which is palinspastically positioned in Figure 2. Section 1 is located at the north end of Takijug Lake, approximately 95 km due northwest of sections 2 and 3. The box marks the location of the map shown in Figure 3. RL, Rockinghorse Lake; PH, Peacock Hills; WC, Wolverine Canyon; HR, Hackett River; WR, Western River.

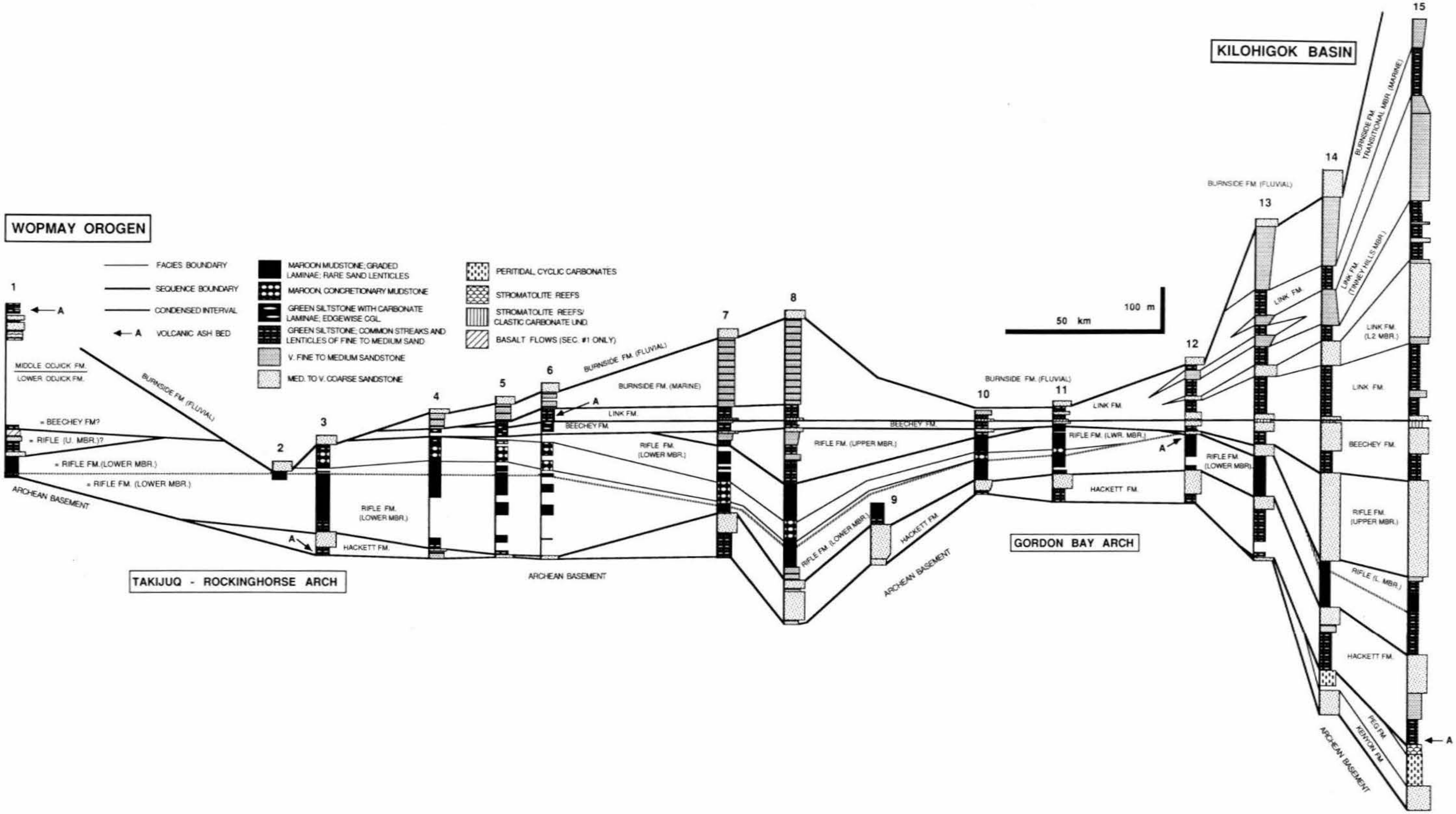

Figure 2. Stratigraphic correlations between Bear Creek Group (Kilohigok Basin; sections 4-15); Rockinghorse Outlier; sections 2,3) and lower Coronation Supergroup (Odjick Formation; Wopmay Orogen; section 1). Section 1 is located at the north end of Takijug Lake, approximately 95 km due northwest of sections 2 and 3.

upward sequence that culminates in a 20-m-thick section of basalt flows. Rapid subsidence is demonstrated by units above this level, consistent with the initiation of thermal (passive) subsidence.

The ramifications of the correlations combined with U-Pb zircon dating are manifold. These involve the interpretation and temporal calibration of major events in both Wopmay Orogen and Thelon Tectonic Zone, the establishment of geodynamic and temporal links between the two, and the quantitative assessment of early Proterozoic lithospheric rheology. Manuscripts are in preparation by J.P. Grotzinger and S.A. Bowring concerning these issues.

SEQUENCE STRATIGRAPHY OF RIFLE FORMATION IN THE TINNEY HILLS AREA

Sequence stratigraphy overview

The concepts of sequence stratigraphy (Vail, 1987) provide a new way of organizing and analyzing stratigraphic and facies data from sedimentary rocks in a chronostratigraphic framework that can be used in strata where no biostratigraphic control is present. Much of the terminology used in this section is defined in Van Wagoner, et al. (1987) and Vail (1987). The following discussion is to be read in conjunction with Figures 3 and 4.

In the Tinney Hills area the Rifle Formation is divided into four sequences with sequence boundary I, the oldest boundary in the Rifle, essentially at the Rifle/Hackett formation boundary and the associated sequence I, encompassing the lower Rifle siltstones and the lower Rifle sandstones (Fig. 3,4). Sequence I has the greatest preserved thickness of sediments in the Rifle Formation and is essentially a complete sequence with lowstand, transgressive and highstand systems tracts represented, though lowstand sands are missing from this area. This relative completeness may reflect an influence on sedimentation controlled primarily by eustatic sea level with any tectonic influence simply augmenting the eustatic signature. In this case the combined signature of eustasy and tectonics is to create accommodation through a relative rise in sea level.

Sequences II and III are wholly within the upper Rifle sandstones and are much thinner than sequence I. Sequence II is made up mostly of highstand deposits with very thin lowstand and transgressive deposits. The bulk of the sandstones in these two systems tracts was probably deposited away from the area of the Tinney Hills outcrops, with the lowstand sediments farther to the south (basinward) and the transgressive sediments farther to the north (toward Gordon Bay Arch). The thin total thickness and the large lateral distances between deposits are probably the result of a stronger tectonic influence on sedimentation interfering with a pure eustatic signature coupled with sedimentation in an area of very low topographic gradient. The effect of the combined tectonic and eustastic signal was to create very little accommodation resulting in thin preserved deposits.

In sequence III the only lowstand sandstones preserved in this area are fluvial sediments filling an incised valley. The incised valley is at least 40 m deep and less than 6 km wide and was cut by sequence boundary III into the highstand units of sequence II during the fall in sea level that created the sequence boundary. Lowstand nearshore marine sandstones are probably located to the south (basinward). The transgressive systems tract has several parasequences but is cut into by sequence boundary IV which has also eroded away all of the sequence III highstand deposits. In this sequence eustasy alone can explain the distribution of lowstand and transgressive deposits and tectonics may have only contributed to the large fall of sea level that created sequence boundary IV.

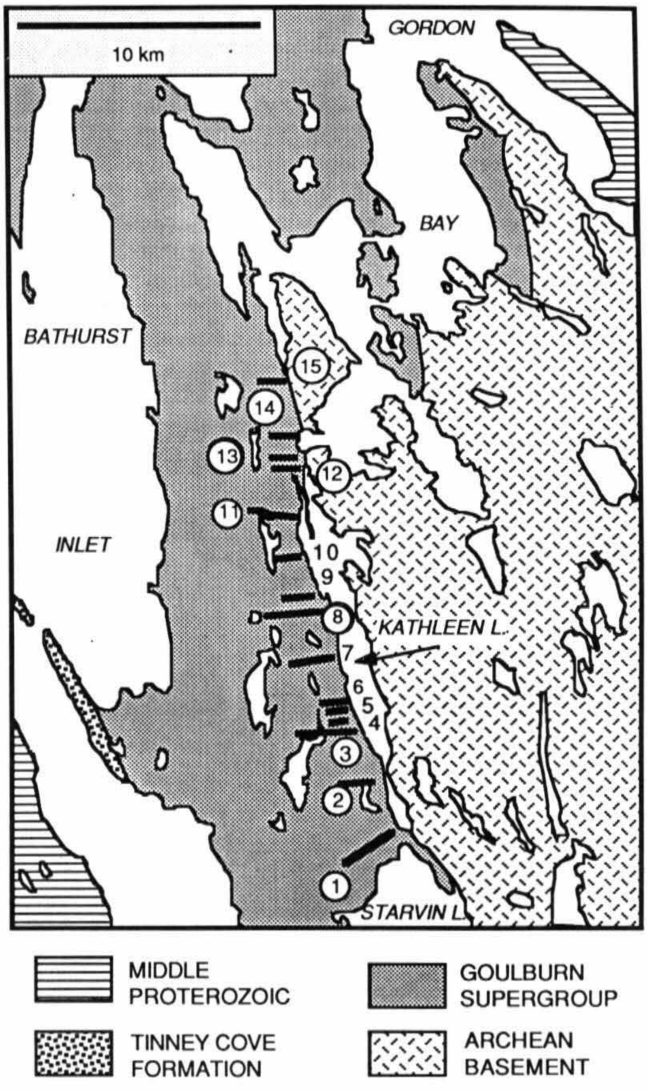

Figure 3. Location map of measured sections illustrated in Figure 4. See Figure 1 for location of this map in Kilohigok Basin.

Most of sequence IV is in the basal part of the Beechey Formation except for thin lowstand and transgressive systems tract deposits within the Rifle in the southern Tinney Hills area. Over most of the Tinney Hills there are no preserved lowstand and transgressive deposits, and the condensed secton rests directly on the sequence boundary. This is the result of a rapid rate of tectonic uplift amplifying a eustatic fall and dominating a eustatic sea level rise with no accommodation being created over much of the Tinney Hills area until the earliest highstand. The highstand siltstones and carbonates of the lower Beechey Formation were not studied for this part of the project and are not discussed.

Sequence descriptions

Each of the four sequences is bounded above and below by sequence boundaries that are Vail unconformities. Each unconformity was formed in response to an abrupt lowering of relative sea level caused by a combination of eustatic sea level change and tectonic uplift. Unconformities in the Rifle Formation can be recognized by a combination of several criteria: a regional surface of exposure with nondeposition and/or erosion; one or more wide spread and well-developed paleosols; an abrupt basinward shift in depositional environments. In the Tinney Hills, erosion is recognized along at least part of each sequence boundary.

Sequence I

The lowest sequence boundary, sequence boundary I, is located at the formation boundary between the Rifle Formation and the underlying Hackett Formation. In several areas there are debris flow deposits directly on top of lower shoreface sandstones of the Hackett Formation and these debris flows are interpreted as forming in response to a rapid fall in sea level. The change in sea level resulted in erosion of more landward sediments that were transported in debris flows to their present position directly on top of the sequence boundary.

Shelf siltstones and mudstones are found on top of these debris flows and also directly on top of the Hackett Formation where debris flows are absent. The fine-grained sediments are distal equivalents of nearshore sandstones of the lowstand systems tract. It is important to note here that there is a general upwards increase in the amount of carbonate concretions in the shales that culminates in a laterally continuous one-to-two metre thick dolomite bed near the middle of the lower Rifle siltstone. The dolomite bed is unusual because of its thickness, lateral continuity, and the presence of small, deep-water stromatolites that colonized areas of hardgrounds that formed during this period of starved deposition. An increase in percent of concretions is also seen in the far southern part of the Tinney Hills in a more distal position from the source of clastic sediments, but there the dolomite bed is found in the lower part of the siltstone (compare section 1 with section 5). The increase in numbers of concretions is most likely due to a decrease in the rate of siliciclastic sedimentation. This trend culminates in the formation of the dolomite bed under conditions of almost no influx of clastic sediments resulting in a starved basin. The dolomite bed is interpreted to be a condensed section on top of lowstand siltstones. During this time clastic sedimentation occured far landward of the Tinney Hills during coastal onlap, resulting in greatly diminished transport of siliciclastic sediment into this part of the basin.

The siltstones above the condensed section generally have few concretions and exhibit an overall coarsening-up trend consisting of increasing numbers and thicknesses of thin (less than 50 cm), wave-influenced, sandstone beds. These siltstones are the distal equivalents of early highstand nearshore marine sandstones. A rapid vertical facies change records the arrival of later highstand nearshore marine sandstones deposited as the shoreline prograded into the area of the Tinney Hills. This sandstone is a major Gilbert-type delta complex and is the final preserved unit in the first sequence.

Sequence II

Sequence boundary II is formed on top of the Gilbert-type delta complex in the highstand of sequence I. North of section 5 it is characterized by an exposed surface with incipient paleosol development. In some areas small-scale (less than one metre of relief) karst developed at the sequence boundary, formed by erosion of carbonate-cemented (micrite) fluvial sandstones. Maximum progradation of the underlying delta complex stopped just south of section 5 as shown by the preservation of depositional relief along the delta-front beds just below the boundary and by knickpoint erosion that removed some delta-front sands and deposited them farther downslope as debris flows. These debris flows can be correlated several kilometres to the south into a part of the section that is dominantly siltstone with only minor amounts of sandstone. The initiation of erosion of delta front beds and resultant debris flows records a fall in sea level and a basinward shift in deposition. The surface formed by this can be correlated landward as a surface of little to no deposition with long exposure and paleosol development. Thin fluvial deposits and paleosols are preserved on top of the sequence boundary.

Little of the lowstand systems tract for sequence II is found in the Tinney Hills. Apparently, most of the lowstand deposition bypassed this area and occurred in a more basinal location. This was probably due to the fall in sea level moving the shoreline basinward of the Tinney Hills area. However, the sequence boundary in the vicinity of sections 5, 4 and 3 does parallel the dipping depositional surface of the underlying delta front and this dip may have prevented building up of any substantial accumulation of sand if the shoreline did remain in the vicinity of the southern Tinney Hills. If this is the case, then the sands may have repeatedly flowed farther downslope as gravity-slide deposits and are probably, in part, equivalent to turbiditic sands present in the Bear Creek Hills area.

The first nearshore sandstone of the transgressive systems tract to be preserved in this area can be seen in sections 4 and 5 and is a small deltaic parasequence. This parasequence onlaps the sequence boundary near the top of the preserved delta front of the underlying highstand deposits and deltaic facies grade into fluvial sediments and associated paleosols. Any succeeding nearshore marine parasequences

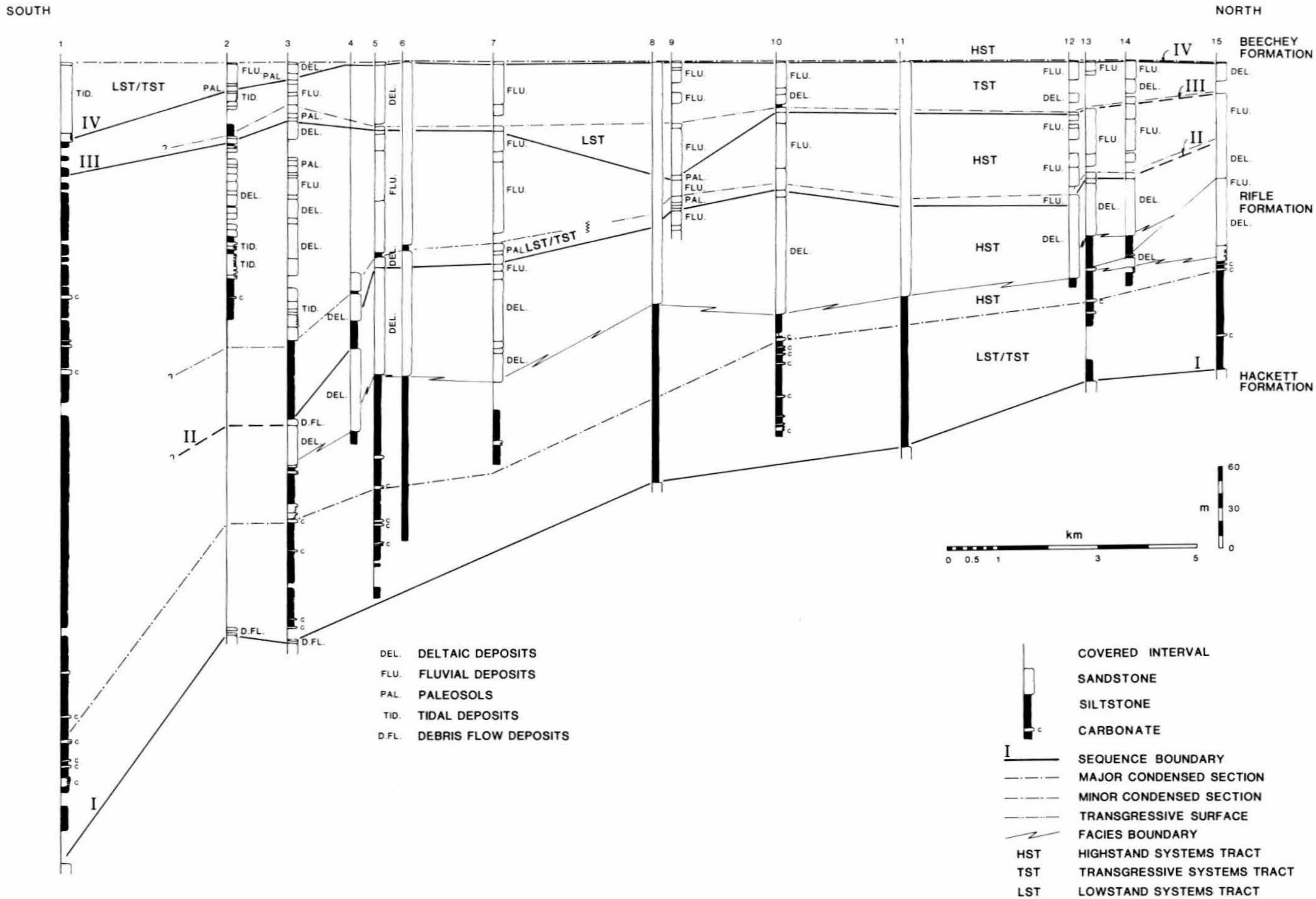

Figure 4. Sequence stratigraphy of Rifle Formation in Tinney Hills area. See text for discussion.

stepped shoreward a great enough distance to be north of the Tinney Hills outcrop. Such a significant backstep indicates that the coastal plain at this time probably was at a very gently inclined angle. This low gradient for the coastal plain is consistent with the apparent very rapid regression of the shoreline as the highstand prograded back over the area. A minor condensed section is preserved in sections 3 through 7 and is interpreted to be the result of starved sedimentation in shallow shelf waters while nearshore sedimentation was occurring farther landward. This supports the interpretation that additional nearshore lowstand parasequences were deposited north of the Tinney Hills. This surface can be correlated across the Tinney Hills north of section 7, shows decreasing influence of submergence to the north, and has less than 1 m of recognizable marine sandstone on top of it.

North of section 5 the only preserved sediments between sequence boundary II and the transgressive surface are fluvial sediments and associated paleosols that were deposited during the transgressive systems tract. It is also likely that some of this thin non-marine section was deposited during part of the lowstand systems tract. Accommodation for non-marine sediments can be low-to-nonexistent during lowstand as it can also be for non-marine sediments associated with initial parasequences of the transgressive systems tract. This explains why there is only a few metres of sandstones assigned to the lowstand and transgressive systems tracts of sequence II.

The highstand systems tract is composed of fluvial sandstones and paleosols from section 6 to the north end of the Tinney Hills; to the south there is an interfingering with nearshore marine units. Both tidal deposits and Gilbert-type delta deposits occur that are dominantly sandstone and which in turn interfinger and grade laterally to the south with deeper-water siltstones. This trend continues until in section 1 only siltstones are preserved for this interval.

Sequence III

The third sequence boundary is very much like the second boundary in that it is a marked surface of exposure and paleosol development. However, erosion into the underlying nearshore marine highstand deposits is not shown in the Tinney Hills outcrops. Instead there is an incised valley formed at the sequence boundary that cuts down at least 40 m (see section 9) into the highstand fluvial deposits of sequence II. The sequence boundary is also a marked surface of exposure and paleosol development at the base of this incised valley implying that there was a period of non-deposition, exposure and weathering before infilling with subsequent fluvial deposits. The incised valley is recorded in section 9, but is not recognized in either section 7 to the south or 10 to the north. This constrains the exposed width of the valley to some value less than the approximately 6 km between sections 10 and 7. The incised valley is a part of the lowstand systems tract of sequence III and will connect downdip with nearshore lowstand deposits: however these are not preserved in the Tinney Hills outcrops and are probably present as part of the basinal turbidite sequence in the Bear Creek Hills area. Between sections 2 and 1 there is a facies change from thin fluvial/paleosol deposits to marine siltstones. The lack of well-developed nearshore marine sandstones in this transition may be due to either a silty and muddy low-energy shoreline at this time or to a facies change from nonmarine to lagoonal or distributary bay facies.

A transgressive surface is recognized on top of the thin fluvial/paleosol deposits and on top of the fluvial deposits in the incised valley. This transgressive surface has also left an overprint on the paleosols beneath it, but to a lesser degree than is found along the transgressive surface in sequence II. The deposits of the transgressive systems tract include a mix of small Gilbert-type deltaic sediments and fluvial units forming at least two parasequences with the line of the outcrop oblique to depositional strike. Tidal deposits of this systems tract are also found in section 2 and these are in a probable third parasequence on top of the deltaic units. There is also a lateral facies change between the tidal sandstones of section 2 and the marine siltstones of 1.

No highstand deposits of sequence III were preserved in the Tinney Hills. All deposits were stripped off by the formation of sequence boundary IV. It is not uncommon to have part or all of the highstand systems tract removed by thesucceeding sequence boundary although this is the first occurrence of this in the Rifle Formation.

Sequence IV

From a little south of section 5 to the north end of the Tinney Hill, sequence boundary IV is immediately overlain by a 1- to 2- m thick mixed siliciclastic and carbonate bed that is the condensed section for sequence IV. No lowstand or transgressive systems tract sediments are preserved between the sequence boundary and the transgressive surface. The formation boundary between the Rifle Formation and the overlying Beechey Formation is placed at the base of the mixed siliciclastic and carbonate bed that forms the condensed section and this places the sequence boundary and the formation boundary at the same surface. Immediately overlying the mixed bed are the Beechey siltstones and carbonates that are a part of the highstand of sequence IV. It is important to note here that the formation boundary between the Rifle and the Beechey formations coincides with the sequence boundary only when the sequence boundary coincides with the base of the condensed section. This contrasts with the first sequence boundary that is essentially equivalent to the Rifle/Hackett formation boundary.

The sequence boundary is an eroded surface with evidence for exposure and weathering and a very strong overprinting by the condensed section. Microdigitate stromatolites are often found within carbonate-filled dissolution vugs in this interval. Just south of section 5 the sequence boundary and the condensed interval start to diverge from each other and lowstand or transgressive systems tract sediments are preserved. It is not possible to resolve which systems tract these sediments belong to due to a lack of sufficient outcrop to determine fully the lateral relationship of these sediments to the sequence boundary. The sandstones at the top of sections 3 and 2 are fluvial whereas the poorly exposed sandstones at the top of section 1 are probably tidal in origin. The thin, mixed siliciclastic and carbonate bed of the condensed section is at the top of the sandstones in all three sections and is in turn overlain by the highstand Beechey siltstones and carbonates. There is about 12 m of fluvial sandstone between sequence boundary and condensed section in section 3, about 22 m of fluvial deposits in section 2, and about 55 m of tidal deposits in section 1. Note again that in this southern part of the area there is a large separation between sequence boundary and formation boundary.

In section 2 the sequence boundary localy scours down more than 8 m into the underlying transgressive systems tract of sequence III. There is evidence of exposure and paleosol development with some fluvial deposition along the southern part of the sequence boundary, especially where the scour and channel-formation occurs. The channel is about 150 m wide across the face of the outcrop.

Depositional environments

Nearshore marine depositional environments in the upper Rifle Formation are interpreted to include tidal and Gilbert-type deltaic environments. Deltaic deposits occur more frequently than do the tidal deposits and also make up the thickest individual units. Additionally, deltaic deposits are found in the northern and central part of the area studied whereas tidal deposits occur in the southern part (more basinward). Two varieties of sandy, braided to low-sinuosity fluvial systems are recognized; one is thought to be more distal to source whereas the other is more proximal. A spectrum of paleosols can also be recognized which are critical to identifying exposure surfaces along sequence boundaries.

Deltaic units

The deposits interpreted to be formed in a type of Gilbert delta are characterized by pronounced sloping foreset surfaces (clinoforms) that dip to the south-southwest at 5 to 25 degrees (after correction for structural dip) and are overlain by horizontal topset beds. The dominant sedimentary structure in these foresets is 2$^1/_2$D crossbedding formed from straight-crested dunes with scour in front of the dune. Preserved height of the crossbedding ranges from 5 to 120 cm with most in the range of 15 to 50 cm. Trough (3D) and planar (2D) cross-bedding occur but are secondary and tend to be smaller; seldom with a preserved height greater than 60 cm and most in the range of 5 to 40cm. Crossbeds are arranged in tabular to wedge-shaped bedsets with gently undulatory to planar erosional bedset surfaces. Broad, shallow channels from 1 to 5 m deep and 20 to 60 m wide are common in the foreset portion of the deltaic units, especially in the more proximal foresets. These channels are filled with 21/2D, trough, and planar crossbedding organized in bedsets that are oriented as lateral accretion deposits. Foreset sandstone is subarkosic with medium to coarse, subround to subangular grains with quartz and carbonate cements. Carbonate cementation promotes pitted weathering surfaces.

The topset part of the Gilbert-type deltaic deposits is also comprised of crossbedding, with the occurrence of trough and 2$^1/_2$D crossbeds about equal in frequency. Preserved height ranges from 10 cm to 40 cm with occasional 50- to 60- cm beds; bedset geometry is similar to that in the foresets but channels tend to be uncommon. Topset sandstone is also subarkosic, and generally medium to coarse, but coarse to very coarse sand and micrite-cemented sandstone clasts 1 to 10 cm are also common. Subangular to subround grains are both quartz and carbonate cemented and intervals with carbonate cements and pitted weathering occur frequently.

The most distal sandstone deposits are referred to as 'toeset' units. The transition from prodelta siltstones and shales to toeset sandstones is very rapid, occurring over a vertical interval of a few tens of centimeters. Toeset sedimentary structures include: horizontal, planar-laminated beds of sandstone, siltsone and shale from 1 to 10 cm thick; current-rippled beds of sandstone and siltstone from 1 to 10 cm thick; and 2$^1/_2$D crossbeds of sandstone from 5 to 35 cm thick. Bedsets are tabular with little, if any, erosion at the base. Planar-laminated sandstones are very fine to fine; current ripples are medium to fine; and 2$^1/_2$D crossbeds are medium to coarse depending on,the height of the crossbed, smaller beds having finer grains. Composition is subarkosic to arenitic with quartz cement.

Within the toesets there are vertical trends of coarsening up, of an increase in bed thickness, and a change from mostly planar-laminated beds to current ripples to crossbeds. Most of the grain-size segregation in the delta occurs in the toeset beds with little segregation in the foresets, even though they make up the bulk of the delta, and only some segregation in the topset beds. There is very little silt- and clay-sized material in the toeset, foreset, or topset parts of the delta deposits and this is interpreted to be due to efficient bypassing of these sub-environments with deposition of fines in the prodelta/open shelf area.

Overall trends of the Gilbert-type deltas include coarsening up, thickening up of beds, a progressive change to higher energy sedimentary structures, and the gross change in bedset geometry from toeset to foreset to topset.

Tidal deposits

Tidal deposits, found in the southern part of the field area, are characterized by relatively thin, 1- to 5- m, intervals that have pronounced vertical trends of either thickening and coarsening upwards or thinning and fining upwards . A higher percentage of shale and siltstone is found in these deposits than is found in either the Gilbert-type deltaic units or the fluvial deposits described below. Average sizes of bedsets, beds and sedimentary structures are smaller in the tidal deposits than in the deltaic rocks. Tabular bedsets are generally less than 70 cm thick and average 10 to 40 cm thick. Trough, 2$^1/_2$D, and planar crossbeds range up to 80 cm in preserved height and average 20 to 30 cm; current and wave ripples are 1 to 3 cm high; beds of horizontal, planar laminated sandstones, siltstones, and shales are less than 20 cm thick with most less than 10 cm; flaser, wavy, and lenticular bedding are all 5 to 10 cm thick. Subarkosic to arkosic sandstones with very fine to coarse, subround to subangular grains are characteristic of these rocks. The tidal deposits include intervals interpreted as tidal channels, possible tidal deltas, and tidal flats based on the vertical association of sedimentary structures and the vertical trends in bed thickness and grain size.

One 2-m-thick interval with several tidal channel deposits was identified. Each channel is comprised of $2^1/2$D crossbeds arranged in lenticular bedsets nested within a lenticular scour cut into lower tidal flat deposits. Coarse to medium sand is found within the channels and in the surrounding tidal flat beds. These deposits are characterized by their lenticular geometry.

Possible tidal delta deposits are 2 to 5 m thick and are marked by coarsening up and thickening up trends. Sedimentary structures at the base include mostly current ripples with some wave ripples and horizontal planar laminated beds comprised of very fine to medium sand, silt, and clay. The top is made up of $2^1/2$D and planar crossbeds 10 to 30 cm in preserved height arranged in tabular bedsets and is comprised of medium and coarse sand. There is also a gradual transition between upper and lower parts of these deposits with respect to sedimentary structures and grain size.

Intervals 1.5 to 4.5 m thick and interpreted as possible tidal flat deposits have marked fining up and thinning up vertical trends with mostly medium to coarse sand at the base, and fine to medium sand with interbedded and interlaminated silt and clay at the top. Trough and $2^1/2$D crossbeds are found in the base, often with reactivation surfaces that form sigmoidal bedsets (bundles). Current ripples, flaser bedding, wavy bedding, lenticular bedding, and horizontal planar laminated beds are all found the upper portion. There is a gradual transition between upper and lower parts both in dominant grain size and in types of sedimentary structures.

Fluvial deposits

Two distinct styles of sand-rich fluvial deposition are recognized in the Rifle Formation. One style is interpreted to be somewhat proximal, sand-rich, braided river deposits whereas the second is interpreted to be either more distal braided river deposits or low-sinuosity fluvial deposits. The proximal braided river sandstones are thought to have been deposited in a braid plain environment and the distal fluvial sandstones are thought to have been deposited in a coastal plain environment close to the shoreline. Proximal fluvial sandstones are characterized by a lack of tabular bedsets, trough crossbedding with a subordinate amount of $2^1/2$D crossbedding, a reddish-brown weathering color, and a large number of intervals with carbonate (micrite) cement and very pitted weathering surfaces. Distal fluvial sandstones are characterized by tabular bedsets, roughly equal proportions of trough and $2^1/2$D crossbedding, common broad and shallow channels, and interbedded micrite-cemented intervals with most of the micrite cement concentrated in the tops of beds and bedsets. There is little vertical change in grain size, bed or bedset thickness, or types of sedimentary structures in either the proximal or distal fluvial deposits.

Proximal fluvial sandstones

Trough crossbedding predominates in these sandstones with some $2^1/2$D crossbedding also occurring. Preserved height of the crossbedding ranges from 5 to 80 cm, with an average of 15 to 35 cm. There are also some contorted crossbeds scattered throughout the deposits, often at the base of an interval. Grain size varies from medium to very coarse, many of the deposits are coarse to very coarse and overall the proximal deposits are coarser than are the distal deposits. The sandstone is subarkosic in composition and grains are subround to subangular.

Distal fluvial sandstones

Trough and $2^1/2$D crossbedding occur with approximately equal frequency in these sandstones and range in preserved height from 5 to 50 cm with an average of 15 to 40 cm. A few current ripples 2 to 5 cm high occur with small trough crossbeds 5 to 20 cm in height in one section and may represent rarely preserved deposition at or near the top of a bar. Contorted crossbeds are about as common in the distal fluvial sandstones as in the proximal sandstones. Tabular bedsets range from 20 to 150 cm in thickness with most ranging from 50 to 100 cm. Some broad, shallow channels occur and are filled with trough and $2^1/2$D crossbedding. These channels are similar to those described in the foreset and topset beds of the Gilbert-type deltas. Grain size averages medium to coarse but some very coarse sand also occurs. Some sandstone clasts 7 to 15 cm also occur. Composition is subarkosic and grains are subround to subangular.

Topset sandstones of the Gilbert-type deltas are very similar to the distal fluvial sandstones with respect to sedimentary structures, bedset geometry, channels, grain size and composition, and general overall appearance. It is often difficult to decide where topset deposits end and distal fluvial deposits begin and both vertical and lateral position of a unit must be taken into account. It is this strong similarity to topset deposits that has guided the interpretation of these fluvial deposits as being formed in a coastal plain environment distal to a sediment source and close to the shoreline.

Alternating proximal and distal fluvial sandstones

Two intervals in Section 7 (50 m and 20 m in thickness) have alternating proximal and distal fluvial sandstones where the proximal sandstones are less than or equal to 1 m in thickness and the distal sandstones are less than or equal to 2 m in thickness. This thin and repeated alteration may be due to the section being located in an area of interfingering facies or it may record parasequences in a subaerial setting with little accommodation. Each pair or couplet of distal overlain by proximal is capped by carbonate micrite cement in the upper 10 to 30 cm, interpreted to likely represent a period of exposure and nondeposition.

MARINE-ALLUVIAL TRANSITION AND STRATIGRAPHIC RELATIONSHIPS WITH THE LOWER BURNSIDE FORMATION

The 1988 season of sedimentological and stratigraphic field studies of the Burnside Formation concentrated on the transition from the marine shelf facies in the upper Link Formation to the braided alluvial facies which characterize the majority of the Burnside Formation. The details of alluvial facies and overall evolution of the Burnside Formation were discussed in previous reports (Grotzinger et al. 1987; McCormick and Grotzinger 1988). This section outlines two aspects of the marine-alluvial transition. First, the sedimentology of the transition at one superbly-exposed section (north Tinney Hills) is detailed. Second, the stratigraphy of the upper Link and the lower Burnside formations from the Tinney Hills to the crest of the Gordon Bay Arch are shown and the implications are discussed for the transition within the Kilohigok foreland basin from southeast-facing marine shelf ramp to northwest-flowing alluvial system.

Marine-alluvial transition

The marine to alluvial transition in the north Tinney Hills is characterized by three main associations of facies which represent storm-influenced marine shelf, lower delta slope, and upper delta slope.

The marine shelf association comprises six grayish-green siltstone facies which are commonly arranged in the following vertical succession: laminated mudstone, mm-scale graded mudstone-siltstone couplets; millimetre-to centimetre-scale siltstone in mudstone commonly containing starved wave ripples (Fig. 5a); 5-30 cm beds of hummocky or swaley cross-stratified coarse siltstone and very fine-grain sandstone (Fig. 5b); parallel- to wavy-laminated silt and sand with common siltstone- and sandstone-filled gutters 5-20 cm deep (Fig. 5c); and massive to parallel-laminated siltstone beds 5-200 cm thick. This progression of facies suggests deposition on a storm-influenced muddy shelf (Johnson and Baldwin 1986). The upward sequence of textures and sedimentary structures suggest passage from conditions of suspension fall-out and weak density flows below storm wave-base (laminated mudstone, graded couplets) through storm-wave influenced structures of progressively higher energy and sediment supply (wave rippled siltstone, hummocky/swaley beds, gutter siltstones; Fig. 5a-b). This succession is seen particularly well in one section just below the lower contact of the Burnside Formation (Fig. 6: 0-30 metres) and suggests progressive shallowing and progradation of a storm-influenced muddy shelf. Gutters are very uniformly aligned ($111°/291° \pm 13°$, $n=12$). Thick massive to laminated beds of siltstone are interpreted as prodelta deposits of density currents under conditions of high sediment supply (Elliot 1986). This facies commonly passes up to red delta front siltstones.

The transition from hummocky/swaley cross-stratified facies to gutter facies does not match the sequence of Dott and Bourgeois (1982) of storm-dominated progradational shelves in which the thickest bedded swaley cross-stratified facies represent the shallowest shoreface environment. In this study, the predominance of gutters which locally are isolated in siltstone and their occurrence with thin discontinuous silt or sand beds (Fig. 5c) suggest that storm conditions promoted erosion and bypassing along the nearshore sea floor,rather than deposition of amalgamated hummocky or swaley beds. Myrow et al. (1988) describe a similar succession from the late Precambrian-early Cambrian of Newfoundland which they suggest may be more representative of fine-grained storm-influenced shelves.

The lower contact of the Burnside Formation is marked by the transition from shelf to lower delta slope association. The coarsening-upward delta slope association comprises several facies (Fig. 5d-e). The occurrence of this association is restricted to the lower few metres of the Transitional Member of the Burnside Formation (Fig. 6: 40-51 metres). This association is marked by the first occurrence of red siltstone/mudstone beds which are interbedded with wave rippled, very fine-grained sandstone. Upward, siltstone beds (1-5 cm) appear with single-grain horizontal to wavy-parallel laminae composed of fine to medium sand at the base of beds. There is an upward increase in bed thickness. Current, wave, and interference ripples are common. Locally, hummocky cross-stratified sandstone beds occur. The most distinctive feature of this association, other than the red colour, is the presence of abundant soft sediment deformation structures, such as flame, load, and ball-and-pillow structures, convolute laminae, and entire beds which have slumped (Fig. 5d-e; 6: 43-46 m). No desiccation cracks are found, but irregular, disconnected three-armed cracks are locally present which are interpreted as subaqueous shrinkage (synaeresis) cracks.

The abundance of wave-formed structures, such as wave and interference ripples, the presence of synaeresis cracks, and lack of desiccation cracks, suggest deposition in a very shallow subaqueous environment which was not subaerially exposed. Soft-sediment deformation and red colour support high rate of terrigenous input which caused slope over-steepening and promoted slumping. This association is most consistent with the lower slope of a prograding fluvial-dominated delta (Elliot 1986). This interpretation is supported by vertical stratigraphic trends which put this association between prodelta shelf and upper delta front-delta plain associations (Fig. 6).

The lower delta front association passes upward into upper delta front-delta plain associations which have been described by McCormick and Grotzinger (1988). The upper delta slope comprises mostly tabular-bedded, horizontally-laminated, very fine- to fine-grain sandstone with wave ripples, mud clasts, and internal scouring (Fig. 6: 52-95 m). Passage from delta slope to delta plain is characterized by a change from isolated to laterally and vertically-persistent trough crossbedded fine- to medium-grain sandstone (Figure 6: 95-113 metres). As reported previously, the horizontally-laminated sandstone is interpreted as wave-reworked braid delta distributary mouth shoals and swash bars. The isolated trough crossbeds are interpreted as braid delta distributary channels which occasionally cut across the

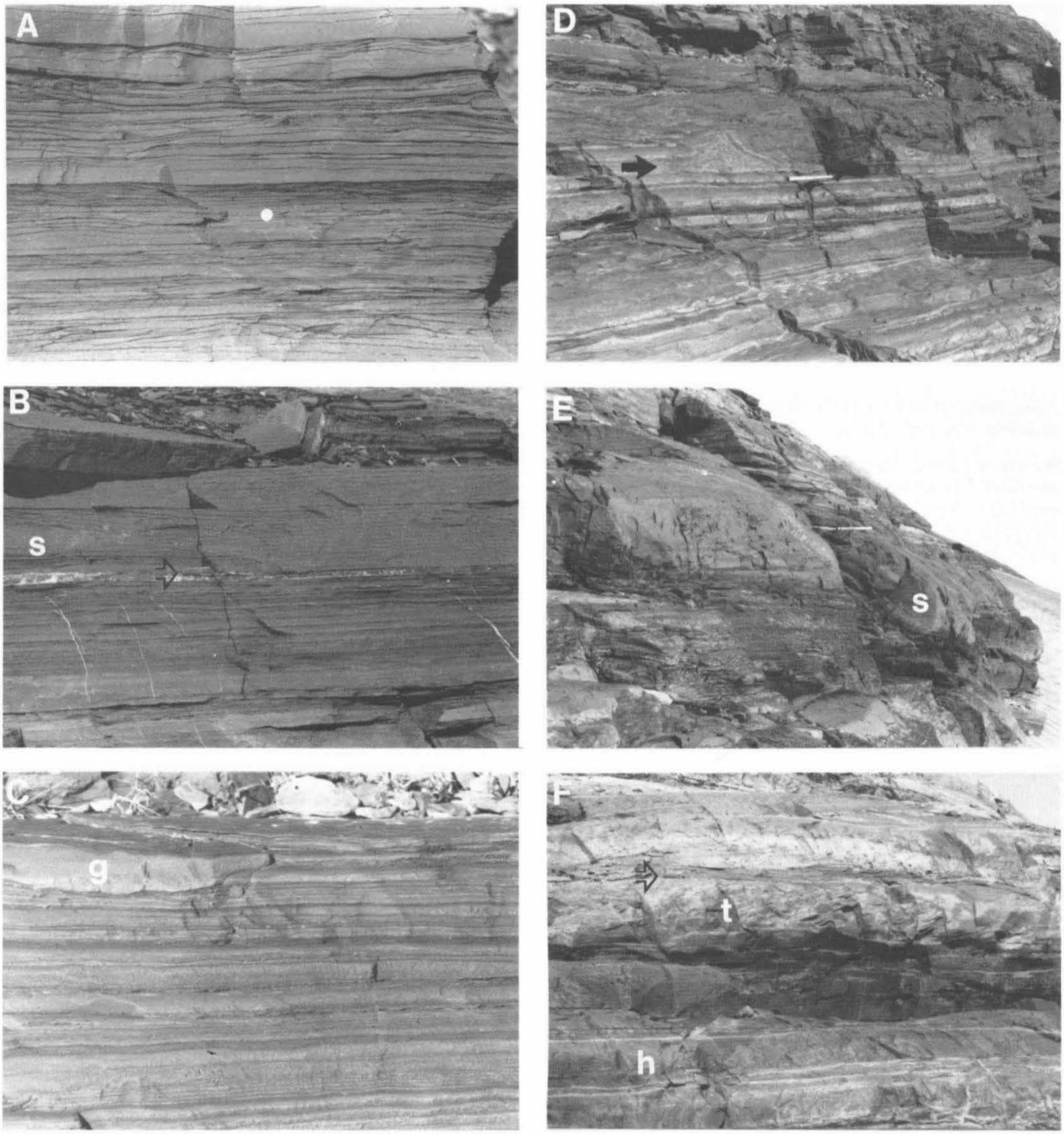

Figure 5. Transitional marine shelf to deltaic associations. A) wave rippled siltstone-mudstone facies. Coin for scale. B) Hummocky-swaley cross-stratified facies. Truncated swaley bed (s) interbedded with wave rippled and horizontally laminated siltstone. Arrow points to pocketknife for scale. C) Gutter facies. Note how gutter (g) passes laterally into thin wavy to parallel laminated siltstone bed. Coin for scale (lower right of gutter). D) Wave rippled red sandstone-siltstone facies. Arrow points to highly deformed horizon which is injected into overlying bed. The overlying bed is highly contorted and passes laterally into a slumped horizon (see next photo). Hammer for scale. E) Slumped bed (s) cuts out about 2.5 m of section. Bed dips at about 30 to regional bedding but has remained internally parallel laminated and coherent. Laterally this bed passes into contorted, fluidized sandstone. Hammer for scale above slumped bed. F) Delta front/delta plain sandstones. Trough cross-bedded sandstone (t) with common mud chips are sharply interbedded tabular horizontally laminated sandstones (h). Arrow points to pocket knife for scale.

delta platform during strong floods. Interestingly, at the section at north Tinney Hills, a strong correspondence between gutter orientations and trough cross bedding in overlying facies at this location suggests that gutters may have been formed by erosive storm-surge ebb currents that were aided by flood currents from distributary channels of the prograding braid delta.

The criteria to distinguish between marine and alluvial facies for the Burnside Formation may be useful for other Precambrian sandstone successions. By distinguishing between the two, it may help to identify unconformity-bounded depositional sequences which are time-stratigraphic units essential for intra-basinal correlation (Christie-Blick et al. 1988; Grotzinger et al. 1988). The identification of such disconformities within the Burnside Formation is discussed below.

Stratigraphic relationships within the Link and lower Burnside formations

The Link and lower Burnside Formations (Transitional and Buff Quartzite members; McCormick and Grotzinger 1988) record the conversion of the Bear Creek foreland basin from a southeast-facing shallow marine ramp to a northwest-flowing alluvial system (Grotzinger et al. 1988). Two major proximal-to-distal trends are evident. First, the lower interval from the base of the Link Formation to the base of the main conglomeratic unit in the Burnside Formation define a northwest-tapering wedge which thins over the Gordon Bay Arch, thickens towards Wolverine Canyon, then thins towards Rockinghorse Lake (Fig. 2). This geometry mimics the underlying lower BearCreek Group (Fig. 2) (Grotzinger and McCormick, 1988). Towards the crest of the Gordon Bay Arch, within the Link Formation, coarsening-upward sequences which culminate in medium-grain, well-sorted, white trough crossbedded sandstone with glauconite are more common. These sequences closely resemble shallow marine sand bars (Johnson and Baldwin, 1986). The increasing abundance of wave-influenced sandy shelf facies in the Link Formation to the northwest suggests that the cratonic foreland experienced shallow water conditions more frequently than the proximal end of the basin. This relationship is predicted for a foreland basin in which lower subsidence rates should exist over the flexural arch where sediment supply rate would be more likely to keep pace with subsidence and thus maintain shallow water conditions.

Within the Transitional Member of the Burnside Formation, delta plain facies pinch out to the northwest and this member is entirely composed of upper delta front facies. The Buff Quartzite Member, which is composed of the medial braid plain facies, pinches out southeast of the arch crest so that the upper delta slope facies is abruptly overlain by proximal braid plain facies (Fig. 2). This suggests that the contact between the Link and Burnside formations is an

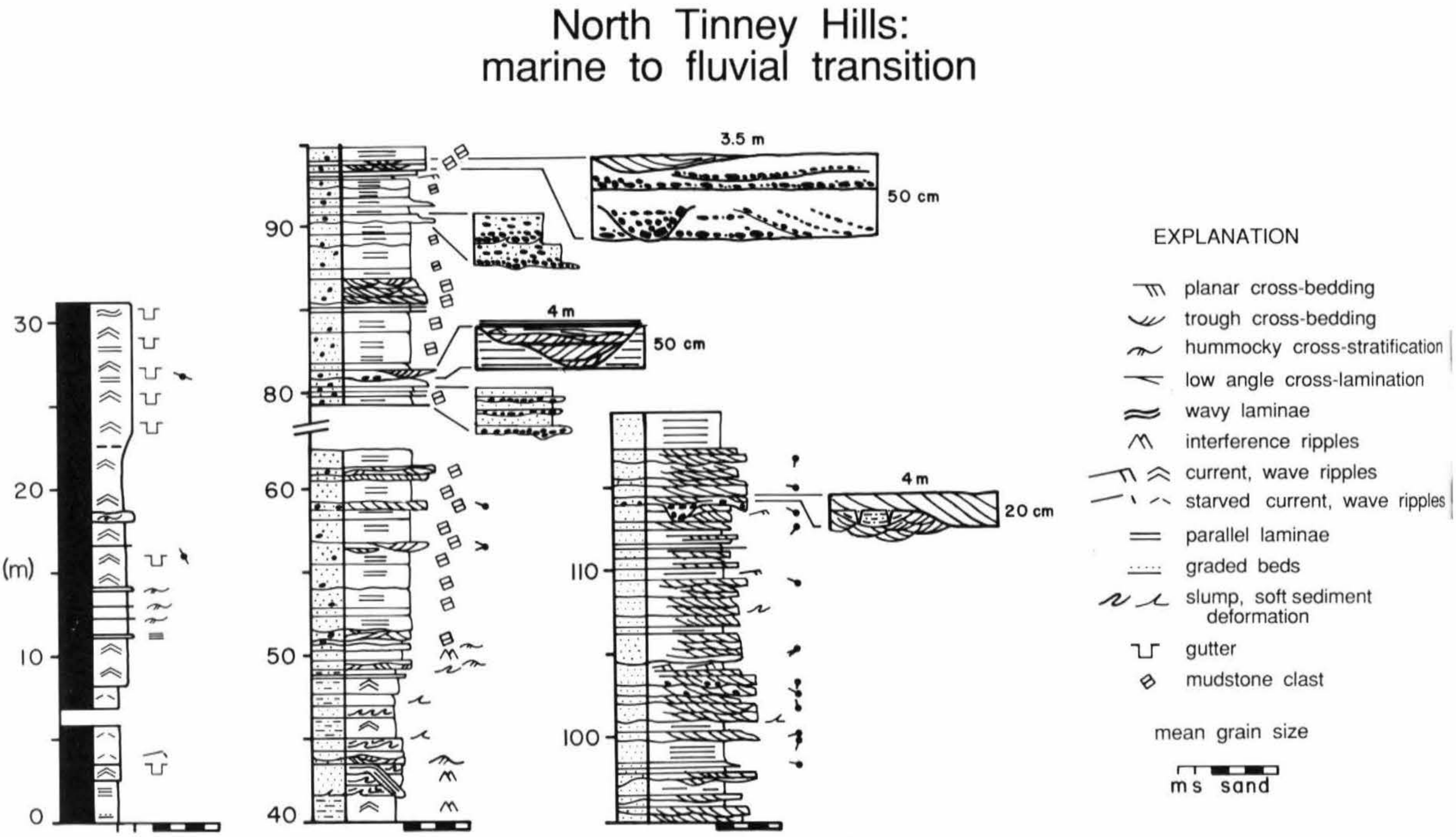

Figure 6. Measured section of the upper Link and Transitional Member of the Burnside Formation at north Tinney Hills. Omitted intervals are covered. Small black circles with tails indicate paleocurrent directions: double tails are gutter trends which are ambiguous with respect to flow direction, and single tails are trough cross-bed azimuths. The northwest to southwest mode of troughs strongly suggest that flow through gutters was to the northwest down a prodelta slope.

erosional unconformity. Northwest of the Gordon Bay Arch, towards Wolverine Canyon this transition is increasingly gradational. At Rockinghorse Lake, the basal Burnside pebbly sandstone rests abruptly on sub-wave-base red mudstones, again suggesting erosional unconformity at the base of the Burnside Formation (Fig. 2).

The transport of gravel across the entire Slave craton, in excess of 200 km, requires a fundamental change in the distribution of subsidence across the basin. Modelling of gravel transport by Paola (1988) suggests that areally-extensive conglomerates, especially in foreland basins, indicate reduced subsidence rates in the proximal part of the basin. Heller et al. (1988) suggest that coarsening-upward profiles in the distal parts of foreland basins indicate when convergence and thrust loading have ceased and erosion predominates which causes unloading and flexural rebound of the hinterland. This uplift causes erosion of previously-deposited foreland basin strata and progradation of the alluvial system over the foreland. This interpretation is supported by uniformly northwest-directed paleocurrents in the foreland indicating long-term overfilling of the depocentre. If this interpretation is correct, then the transition from lower Burnside Formation deltaic and distal alluvial facies to gravelly proximal alluvial facies record this shift from subsidence-dominated foreland sedimentation to erosion- and uplift-dominated sediment redistribution.

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